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Earth Surface Dynamics Discussions

A probabilistic framework for the cover effect in bedrock erosion

Jens M. Turowski
Helmholtzzentrum Potsdam, German Research Centre for Geosciences GFZ, Telegrafenberg, 14473 Potsdam, Germany, turowski@gfz-potsdam.de
Rebecca Hodge
Department of Geography, Durham University, Durham, DH1 3LE, United Kingdom, rebecca.hodge@durham.ac.uk

Abstract
The cover effect in fluvial bedrock erosion is a major control on bedrock channel morphology and long-term channel dynamics. Here, we suggest a probabilistic framework for the description of the cover effect that can be applied to field, laboratory and modelling data and thus allows the comparison of results from different sources. The framework describes the formation of sediment cover as a function of the probability of sediment being deposited on already alleviated areas of the bed. We define benchmark cases and suggest physical interpretations of deviations from these benchmarks. Furthermore, we develop a reach-scale model for sediment transfer in a bedrock channel and use it to clarify the relations between the sediment mass residing on the bed, the exposed bedrock fraction and the transport stage. We derive system time scales and investigate cover response to cyclic perturbations. The model predicts that bedrock channels achieve grade in steady state by adjusting bed cover. Thus, bedrock channels have at least two characteristic time scales of response. Over short time scales, the degree of bed cover is adjusted such that they can just transport the supplied sediment load, while over long time scales, channel morphology evolves such that the bedrock incision rate matches the tectonic uplift or base level lowering rate.

1. Introduction

Bedrock channels are shaped by erosion caused by countless impacts of the sediment particles they carry along their bed (Beer and Turowski, 2015; Cook et al., 2013; Sklar and Dietrich, 2004). There are feedbacks between the evolving channel morphology, the bedload transport, and the hydraulics (e.g., Finnegan et al., 2007; Johnson and Whipple, 2007; Wohl and Ikeda, 1997). Impacting bedload particles driven forward by the fluid forces erode and therefore shape the bedrock bed. In turn, the morphology of the channel determines the pathways of both sediment and water, and sets the stage for the entrainment and deposition of the sediment (Hodge and Hoey, 2016). Sediment particles play a key role in this erosion process; they provide the tools for erosion and also determine where bedrock is exposed such that it can be worn away by impacting particles (Gilbert, 1877; Sklar and Dietrich, 2004).

The importance of the cover effect - that a stationary layer of gravel can shield the bedrock from bedload impacts – has by now been firmly established in a number of field and laboratory studies (e.g., Chatanantavet and Parker, 2008; Finnegan et al., 2007; Hobley et al., 2011; Johnson and Whipple, 2007; Turowski and Rickenmann, 2009; Turowski et al., 2008; Yanites et al., 2011). Sediment cover is generally modelled with generic relationships that predict the decrease of the fraction of exposed bedrock area A^* with the increase of the relative sediment supply Q_s^*, usually defined as the ratio of sediment supply to transport capacity. Based on laboratory experiments and simple modeling, Turowski and Bloem (2016) argued that the focus on covered area is generally justified on the reach scale and that erosion of bedrock under a thin sediment cover can be neglected. However, the behavior of sediment cover under flood conditions is currently unknown and the assumption that the cover distribution at low flow is representative for that at high flow may not be justified (cf. Turowski et al., 2008).

The most commonly used function to describe the cover effect is the linear decline (Sklar and Dietrich, 1998), which is the simplest function connecting the steady state end members of an empty bed when $Q_s^* = 0$ and full cover when $Q_s^* = 1$:

$$A^* = \begin{cases} 1 - Q_s^* & \text{for} \quad Q_s^* < 1 \\ 0 & \text{otherwise} \end{cases}$$
(eq. 1)

In contrast, the exponential cover function arises under the assumption that particle deposition is equally likely for each part of the bed, whether it is covered or not (Turowski et al., 2007).

$$A^* = \begin{cases} \exp(-Q_s^*) & \text{for} \quad Q_s^* < 1 \\ 0 & \text{otherwise} \end{cases}$$
(eq. 2)

Here, exp denotes the natural exponential function.

Hodge and Hoey (2012) obtained both the linear and the exponential functions using a cellular automaton (CA) model that modulated grain entrainment probabilities by the number of neighbouring grains. However, consistent with laboratory flume data, the same model also produced other behaviours under different parameterisations. One alternative behavior is runaway alluviation, which was attributed by Chatanantavet and Parker (2008) to the differing roughness of bedrock and alluvial patches. Due to a decrease in flow velocity, an increase in surface roughness and differing grain geometry, the likelihood of deposition is higher over bed sections covered by alluvium compared to bare bedrock sections (Hodge et al., 2011). This can lead to rapid alluviation of the entire bed once a minimum fraction has been covered. The relationship between sediment flux and cover is also affected by the bedrock morphology; flume experiments have demonstrated that on a non-planar bed the location of sediment cover is driven by bed topography and hydraulics (e.g., Finnegan et al., 2007; Inoue et al., 2014). Johnson and Whipple (2007) found that stable patches of alluvium tended to form in topographic lows such as pot holes and at the bottom of slot canyons, whereas Hodge and Hoey (2016) found that local flow velocity also controls sediment cover location.

The relationship between roughness, bed cover and incision was explored in a number of recent numerical modeling studies. Nelson and Seminara (2011, 2012) were one of the first to model the impact that the differing roughness of bedrock and alluvial areas has on sediment patch stability. Zhang et al. (2014) formulated a macro-roughness cover model, in which sediment cover is related to the ratio of sediment thickness to bedrock macro-roughness. Aubert et al. (2016) directly simulated the dynamics of particles in a turbulent flow and obtained both linear and exponential cover functions. Johnson (2014) linked erosion and cover to bed roughness in a reach-scale model. Using a model formulation similar to that of Nelson and Seminara (2011), Inoue et al. (2016) reproduced bar formation and sediment dynamics in bedrock channels. All of these studies used slightly different approaches and mathematical formulations to describe alluvial cover, making a direct comparison difficult.

Over time scales including multiple floods, the variability in sediment supply is also important. Lague (2010) used a model formulation in which cover was written as a function of the average sediment

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depth to upscale daily incision processes to long time scales. He found that over the long term, cover dynamics are largely independent of the precise formulation at the process scale and are rather controlled by the magnitude-frequency distribution of discharge and sediment supply. Using the CA model of Hodge and Hoey (2012), Hodge (in press) found that, when sediment supply was very variable, sediment cover was primarily determined by the recent history of sediment supply, rather than the relationships identified under constant sediment fluxes.

So far, it has been somewhat difficult to compare and discuss the different cover functions obtained from theoretical considerations, numerical models, and experiments, since a unifying framework and clear benchmark cases have been missing. Here, we propose such a framework, and develop type cases linked to physical considerations of the flow hydraulics and sediment erosion and deposition. We show how this framework can be applied to data from a published model (Hodge and Hoey, 2012). Furthermore, we develop a reach-scale erosion-deposition model that allows the dynamic modeling of cover and prediction of steady states. Thus, we clarify the relationship between cover, deposited mass and relative sediment supply. As part of this model framework we investigate the response time of a channel to a change in sediment input, which we illustrate using data from a natural channel.

2. A probabilistic framework

2.1. Development

Here we build on the arguments put forward by Turowski et al. (2007) and Turowski (2009). Consider a bedrock bed on which sediment particles are distributed. We can view the deposition of each particle as a random process, and each area element on the bed surface can be assigned a probability for the deposition of a particle. When assuming that a given number of particles are distributed on the bed, the mean behavior of the exposed area can be calculated from the following equation:

$$dA^* = -P(A^*, M_s^*, \dots)dM_s^*$$

(eq. 3)

Here, P is the probability that a given particle is deposited on the exposed part of the bed, which may be a function of the fraction of exposed area, the relative sediment supply, the bed topography and roughness, the particle size, the local hydraulics or other control variables. M_s^* is a dimensionless mass equal to the total mass of the particles residing on the bed per area, which is suitably normalized. A suitable mass for normalization is the minimum mass required to cover a unit area, M_0, as will become clear later. The minus sign is introduced because the fraction of the exposed area reduces as M_s^* increases. Similar to eq. (3), the equation for the fraction of covered area A_c^* = 1-A^* can be written as:

$$dA_c^* = P(A^*, M_s^*, \dots)dM_s^*$$

(eq. 4)

As most previous relationships are expressed in terms of Q_s^*, the relation of M_s^* to Q_s^* will be discussed later.

We can make some general statements about P. First, P is defined for the range $0 \leq A^* \leq 1$ and undefined elsewhere. Second, P takes values between zero and one for $0 \leq A^* \leq 1$. Third, $P(A^*=0) = 0$ and $P(A^*=1) = 1$. Note that P is not a distribution function and therefore does not need to integrate to one. Neither does it have to be continuous and differentiable everywhere.

For purpose of illustration, we will next discuss two simple forms of the probability function P that lead to the linear and exponential forms of the cover effect, respectively. First, consider the case that

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all particles are always deposited on exposed bedrock. In this case, formally, to keep with the conditions stated above, we define P = 1 for 0 < A^* ≤ 1 and P = 0 for A^* = 0. Thus, we can write

$$dA^* = -dM_s^* \quad \text{for} \quad 0 < A^* \leq 1$$
$$dA^* = 0 \quad \text{for} \quad A^* = 0$$

(eq. 5)

Integrating, we obtain:

$$A^* = -M_s^* + C$$

(eq. 6)

where the constant of integration C is found to equal one by using the condition $A^*(M_s^*=0) = 1$. Thus, we obtain the linear cover function of eq. (1). Note that the linear cover function gives a theoretical lower bound for the amount of cover: it arises when all available sediment always falls on uncovered ground, and thus no additional sediment is available that could facilitate quicker alluviation. In essence, this is a mass conservation argument. Now it is obvious why M_0 is a convenient way to normalize: in plots of A^* against M_s^*, we obtain a triangular region bounded by the points [0,1], [0,0] and [1,0] in which the cover function cannot run (Fig. 1).

Similarly to above, if we set P to a constant value smaller than one for 0 < A^* ≤ 1, k, we obtain

$$A^* = 1 - kM_s^*$$

(eq. 7)

It is clear that the assumption of $P = k$ is physically unrealistic, because it implies that the probability of deposition on exposed ground is independent of the amount of uncovered bedrock. Especially when A^* is close to zero, it seems unlikely that, say, always 90% of the sediment falls on uncovered ground. A more realistic assumption is that the probability of deposition on uncovered ground is independent of location and other possible controls, but is equal to the fraction of exposed area, i.e., $P = A^*$. In a probabilistic sense, this is also the simplest plausible assumption one can make. Then

$$dA^* = -A^* dM_s^*$$

(eq. 8)

giving upon integration

$$A^* = e^{-M_s^*}$$

(eq. 9)

The argument used here to obtain the exponential cover effect in eq. (9) essentially corresponds to the one given by Turowski et al. (2007). Since this case presents the simplest plausible assumption, we will use it as a benchmark case, to which we will compare other possible functional forms of P.

In principle, the probability function P can be varied to account for various processes that make deposition more likely either on already covered ground by decreasing P for the appropriate range of A^* from the benchmark case $P = A^*$, or on uncovered ground by increasing P from the benchmark case $P = A^*$. As has been identified previously (Chatanantavet and Parker, 2008; Hodge and Hoey 2012), roughness feedbacks to the flow can cause either case depending on whether subsequent deposition is adjacent to or on top of existing sediment patches. In the former case, particles residing on an otherwise bare bedrock bed act as obstacles for moving particles, and create a low-velocity wake zone in the downstream direction. In addition, particles residing on other single particles are unstable and stacks of particles are unlikely. Hence, newly arriving particles tend to deposit either upstream or downstream of stationary particles and the probability is generally higher for deposition on uncovered ground than in the benchmark case. In the latter case, larger patches of stationary particles increase the surface roughness of the bed, thus decreasing the local flow velocity and

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stresses, making deposition on the patch more likely. In this way, the probability of deposition on already covered bed is increased in comparison to the benchmark case.

A simple functional form that can be used to take into account either one of these two effects is a power law dependence of P on A^*, taking the form $P = A^{*\alpha}$ (Fig. 1A). Then, the cover function becomes

$$A^* = (1-(1-\alpha)M_s^*)^{\frac{1}{1-\alpha}}$$

(eq. 10)

Here, the probability of deposition on uncovered ground is increased in comparison to the benchmark exponential case if $0 < \alpha < 1$, and decreased if $\alpha > 1$.

A convenient and flexible way to parameterize $P(A^*)$ in general is the cumulative version of the Beta distribution, given by:

$$P(A^*) = B(A^*; a, b)$$

(eq. 11)

Here, $B(A^*;a,b)$ is the regularized incomplete Beta function with two shape parameters a and b, which are both real positive numbers, defined by:

$$B(A^*; a, b) = \frac{\int_0^{A^*} y^{a-1}(1-y)^{b-1}dy}{\int_0^1 y^{a-1}(1-y)^{b-1}dy}$$

(eq. 12)

Here, y is a dummy variable. With suitable choices for a and b, cover functions resembling the exponential (a=b=1), the linear form (a=0, b>0), and the power law form (a>>b or a<<b) can be retrieved. Wavy functions are also a possibility (Fig. 2), thus both of the roughness effects described above can be modelled in a single scenario. Unfortunately, the integral necessary to obtain $A^*(M_s^*)$ does not give a closed-form analytical solution and needs to be computed numerically.

In principle, a suitable function P could also be defined to account for the influence of bed topography on sediment deposition. Such a function is likely dependent on the details of the particular bed, hydraulics and sediment flow paths in a complex way and needs to be mapped out experimentally.

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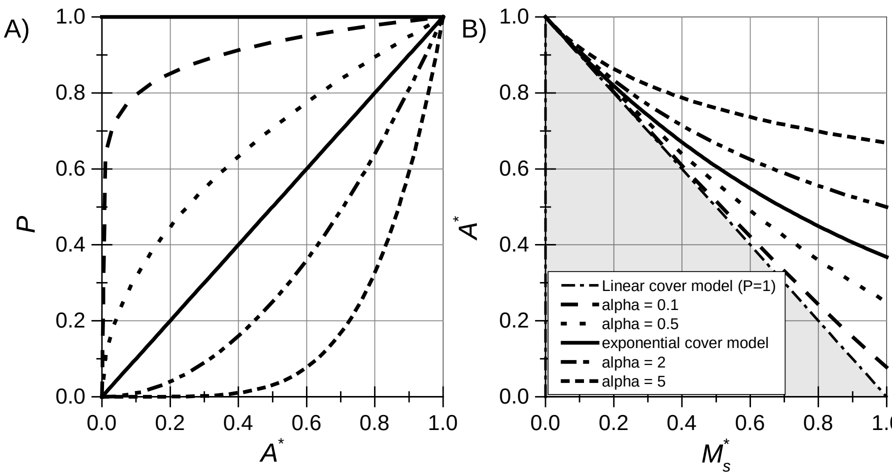

Fig. 1: A) Various examples for the probability function P as a function of bedrock exposure A^*. B) Corresponding analytical solutions for the cover function between A^* and dimensionless sediment mass M_s^* using eq. (7), (9) and (10). Grey shading depicts the area where the cover function cannot run due to conservation of mass.

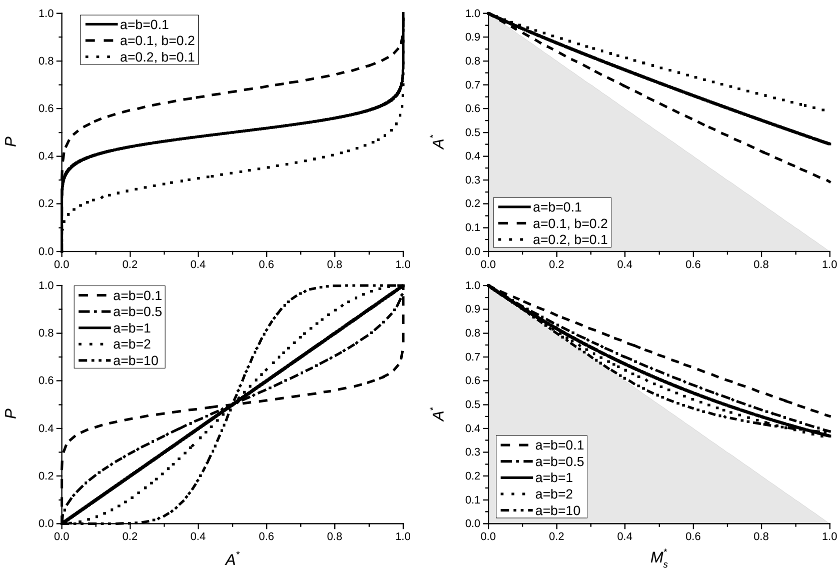

Fig. 2: Examples for the use of the regularized incomplete Beta function (eq. 12) to parameterize P, using various values for the shape parameters a and b. The choice $a = b = 1$ gives a dependence that is equivalent to the exponential cover function. Grey shading depicts the area where the cover function cannot run due to conservation of mass.

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2.2 Example of application using model data

To illustrate how the framework can be used, we apply it to data obtained from the CA model developed by Hodge and Hoey (2012). The CA model reproduces the transport of individual sediment grains over a bedrock surface. In each time step, the probability of a grain being entrained is a function of the number of neighboring grains. If five or more of the eight neighbouring cells contain grains then the grain has probability of entrainment P_c, otherwise it has probability P_i. In most model runs P_c is less than P_i, thus accounting for the impact of sediment cover in decreasing local shear stress (though increased flow resistance) and increasing the critical entrainment shear stress for grains (via lower grain exposure and increased pivot angles).

The model is run with a domain that is 100 cells wide by 1000 cells long, with each cell having the same area as a grain. Up to four grains can potentially be entrained from each cell in a time step, limiting the maximum sediment flux. In each time step random numbers and the probabilities are used to select the grains that are entrained, which are then moved a step length downstream. A fixed number of grains are also supplied to the upstream end of the model domain. A smoothing algorithm is applied to prevent local excessively tall piles of grains. After around 500 time steps the model typically reaches a steady state condition in which the number of grains supplied to and leaving the model domain are equal. Sediment cover is measured in a downstream area of the model domain and is defined as grains that are not entrained in a given time step.

Model runs were completed with a six different combinations of P_i and P_c: 0.95/0.95, 0.95/0.75, 0.75/0.10, 0.75/0.30, 0.30/0.30 and 0.95/0.05. These combinations were selected to cover the range of relationships between Q_s* and A_c* observed by Hodge and Hoey (2012). For each pair of P_i and P_c model runs were completed at least 20 different values of Q_s* in order to quantify the model behaviour.

Cover bed fraction and total mass on the bed given out by the model were converted using eq. (3) into the probabilistic framework (Fig. 3). The derivative was approximated by simple linear finite differences, which, in the case of run-away alluviation, resulted in a non-continuous curve due to large gradients. The exponential benchmark (eq. 9) is also shown for comparison. The different model parameterisations produce results in which the probability of deposition on bedrock is both more and less likely than in the baseline case, with some runs showing both behaviours. Cases where the probability is more than the baseline case (i.e. grains are more likely to fall on uncovered areas) are associated with runs in which grains in clusters are relatively immobile. These runs are likely to be particularly affected by the smoothing algorithm that acts to move sediment from alluviated to bedrock areas. All model parameterisations predict greater bed exposure for a given normalised mass than is predicted by a linear cover relationship (Figure 3b). Runs with relatively more immobile cluster grains have a lower exposed fraction for the same normalised mass. Runs with low values of P_i and P_c seem to lead to behavior in which cover is more likely than in the exponential benchmark, while for high values, it is less likely. However, there are complex interactions and general statements cannot be made straightforwardly.

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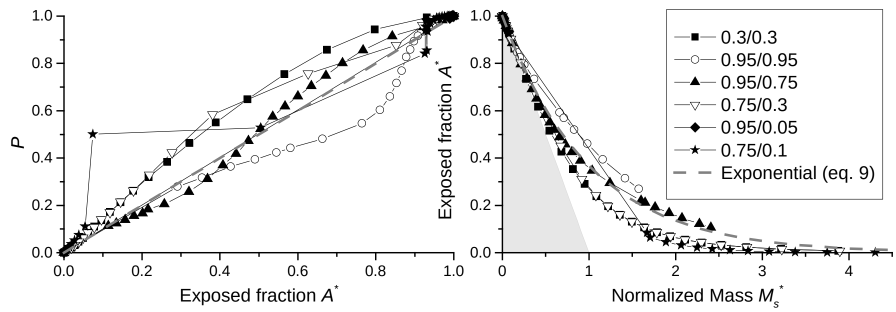

Fig. 3: Probability functions P and cover function derived from data obtained from the model of Hodge and Hoey (2012). The grey dashed line shows the exponential benchmark behavior. Grey shading depicts the area where the cover function cannot run due to conservation of mass. The legend gives values of P_i and P_c used for the runs (see text).

3. Cover development in time and space

3.1. Model derivation

The probabilistic formulation introduced above can be extended to allow the calculation of the temporal and spatial evolution of sediment cover in a stream. Here, we will derive the equations for the one dimensional case (linear flume), but extensions to higher dimensions are possible in principle. The derivation is inspired by the erosion-deposition framework (e.g. Charru et al., 2004), with some necessary adaptions to make it suitable for channels with partial sediment cover. In our system, we consider two separate mass reservoirs within a control volume. The first reservoir contains all particles in motion, the total mass per bed area of which is denoted by M_m, while the second reservoir contains all particles that are stationary on the bed, the total mass per bed area of which is denoted by M_s. We need then three further equations, one to connect the rate of change of mobile mass to the sediment flux in the flume, one to govern the exchange of particles between the two reservoirs, and one to describe how sediment transport rate is related to the mobile mass. The first of these is of course the Exner equation of sediment continuity (e.g. Paola and Voller, 2005), which captures mass conservation in the system. Instead of the common approach tracking the height of the sediment over a reference level, we use the total sediment mass on the bed as a variable, giving

$$\frac{\partial M_m}{\partial t} = -\frac{\partial q_s}{\partial x} + E - D$$

(eq. 13)

Here, x is the coordinate in the streamwise direction, t the time, q_s the sediment mass transport rate per unit width, while E is the mass entrainment rate per bed area and D is the mass deposition rate per bed area. It is clear that for the problem at hand the choice of total mass or volume as a variable to track the amount of sediment in the reach of interest is preferable to the height of the alluvial cover, since necessarily, when cover is patchy, the height of the alluvium varies across the bed. It is useful to work with dimensionless variables by defining $t^* = t/T$ and $x^* = x/L$, where T and L are suitable time and length scales, respectively. The dimensionless mobile mass per bed area M_m^* is equal to M_m/M_0, and eq. (13) becomes:

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$$\frac{\partial M_m^*}{\partial t^*} = -\frac{\partial q_s^*}{\partial x^*} + E^* - D^*$$

(eq. 14)

Here,

$$q_s^* = \frac{T}{LM_0} q_s$$

(eq. 15)

The dimensionless entrainment and deposition rates, E^* and D^*, are equal to TE/M_0 and TD/M_0, respectively. The rate of change of the stationary sediment mass M_s in time is the difference of the deposition rate D and the entrainment rate E.

$$\frac{\partial M_s}{\partial t} = D - E$$

(eq. 16)

Or, using dimensionless variables

$$\frac{\partial M_s^*}{\partial t^*} = D^* - E^*$$

(eq. 17)

We also need sediment entrainment and deposition functions. The entrainment rate needs to be modulated by the availability of sediment on the bed. If M_s^* is equal to zero, no material can be entrained. A plausible assumption is that the maximal entrainment rate, E^*_{max}, is equal to the transport capacity.

$$E_{max}^* = q_t^*$$

(eq. 18)

Here, q_t^* is the dimensionless mass transport capacity, which is related to the transport capacity per unit width q_t by a relation similar to eq. (15). To first order, the rate of change in entrainment rate, dE, is proportional to the difference of E_{max} and E, and to the rate of change in mass on the bed.

$$dE^* = (E_{max}^* - E^*)dM_s^* = (q_t^* - E^*)dM_s^*$$

(eq. 19)

Integrating, we obtain

$$E^* = E_{max}^*\left(1 - e^{-M_s^*}\right) = \left(1 - e^{-M_s^*}\right)q_t^*$$

(eq. 20)

Here, we used the condition $E^*(0) = 0$ to fix the integration constant to E^*_{max}. As required, eq. (20) approaches E^*_{max} as M_s^* goes to infinity, and is equal to zero when M_s^* is equal to zero. Using a similar line of argument, and by assuming the maximum deposition rate to be equal to q_s^*, we arrive at an equation for the deposition rate D^*.

$$D^* = \left(1 - e^{-M_m^*}\right)q_s^*$$

(eq. 21)

Substituting eqs. (20) and (21) into eq. (17), we obtain:

$$\frac{\partial M_s^*(x^*,t^*)}{\partial t^*} = D^* - E^* = \left(1 - e^{-M_m^*(x^*,t^*)}\right)q_s^*(x^*,t^*) - \left(1 - e^{-M_s^*(x^*,t^*)}\right)q_t^*(x^*,t^*)$$

(eq. 22)

Note that $q_s^*/q_t^* = Q_s^*$. The equation for the mobile mass (eq. 14) becomes:

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$$\frac{\partial M_m^*(x^*,t^*)}{\partial t^*} = -\frac{\partial q_s^*}{\partial x^*} - \left(1 - e^{-M_m^*(x^*,t^*)}\right) q_s^*(x^*,t^*) + \left(1 - e^{-M_s^*(x^*,t^*)}\right) q_t^*(x^*,t^*)$$

(eq. 23)

Finally, the sediment transport rate needs to be proportional to the mobile sediment mass times the downstream sediment speed U, and we can write

$$q_s^*(x^*,t^*) = U^*(x^*,t^*) M_m^*(x^*,t^*)$$

(eq. 24)

Here

$$U^* = \frac{T}{L} U$$

(eq. 25)

After incorporating the original equation between A^* and M_s^* (eq. 3), the system of four differential equations (3), (22), (23) and (24) contains four unknowns: the downstream gradient in the sediment transport rate $\partial q_s^*/\partial x^*$, the exposed fraction of the bed A^*, the non-dimensional stationary mass M_s^*, and the non-dimensional mobile mass M_m^*, while the non-dimensional transport capacity q_t^* and the non-dimensional downstream sediment speed U^* are input variables, and P is a externally specified function. In addition, sediment input needs to be specified as an upstream boundary condition and initial values for the mobile and stationary masses need to be specified everywhere.

3.2. Time-independent solution

Setting the time derivatives to zero, we obtain a time-independent solution, which links the exposed area directly to the ratio of sediment transport rate to transport capacity. From eq. (23) it follows that in this case, the entrainment rate is equal to the deposition rate and we obtain

$$\left(1 - e^{-\overline{M_m^*}}\right) \overline{q_s^*} = \left(1 - e^{-\overline{M_s^*}}\right) q_t^*$$

(eq. 26)

Here, the bar over the variables denotes their steady state value. Substituting eq. (24) to eliminate $\overline{M_m^*}$ and solving for $\overline{M_s^*}$ gives

$$\overline{M_s^*} = -\ln\left\{1 - \left(1 - e^{-\overline{q_s^*}/U^*}\right)\frac{\overline{q_s^*}}{q_t^*}\right\} = -\ln\left\{1 - \left(1 - e^{-\frac{q_t^*}{U^*}\overline{Q_s^*}}\right)\overline{Q_s^*}\right\}$$

(eq. 27)

Note that we assume here that sediment cover is only dependent on the stationary sediment mass on the bed and we thus neglect grain-grain interactions known as the dynamic cover (Turowski et al., 2007). In analogy to eq. (24), we can write

$$q_t^* = U^* M_0^*$$

(eq. 28)

Here, M_0^* is a characteristic dimensionless mass that depends on hydraulics and therefore implicitly on transport capacity (which is independent of and should not be confused with the minimum mass necessary to fully cover the bed M_0). When sediment transport rate equals transport capacity, then M_0^* is equal to the mobile mass of sediment normalized by the reference mass M_0. It can be viewed as a proxy for the transport capacity and is a convenient parameter to simplify the equations. The mobile mass can then, in general, be written as (cf. Turowski et al., 2007), remembering that Q_s^* = 1 when transport is equal to capacity:

$$M_m^* = M_0^* Q_s^*$$

(eq. 29)

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If we use the exponential cover function (eq. 9) with eqs. (27), (28) and (29) we obtain

$$\overline{A^*} = 1 - \left(1 - e^{-\overline{q_s^*}/U^*}\right)\frac{\overline{q_s^*}}{q_t^*} = 1 - \left(1 - e^{-\frac{q_t^*}{U^*}\overline{Q_s^*}}\right)\overline{Q_s^*} = 1 - \left(1 - e^{-M_0^*\overline{Q_s^*}}\right)\overline{Q_s^*}$$

(eq. 30)

Similarly, equations can be found for the other analytical solutions of the cover function. For the linear case (eq. 7), we obtain:

$$\overline{A^*} = 1 + \ln\left\{1 - \left(1 - e^{-M_0^*\overline{Q_s^*}}\right)\overline{Q_s^*}\right\}$$

(eq. 31)

For the power law case (eq. 10), we obtain:

$$\overline{A^*} = \left[1 + (1-\alpha)\ln\left\{1 - \left(1 - e^{-M_0^*\overline{Q_s^*}}\right)\overline{Q_s^*}\right\}\right]^{\frac{1}{1-\alpha}}$$

(eq. 32)

It is interesting that the assumption of an exponential cover function essentially leads to a combined linear and exponential relation between $\overline{A^*}$ and $\overline{Q_s^*}$. Instead of a linear decline as the original linear cover model, or a concave-up relationship as the original exponential model, the function is convex-up for all solutions (Fig. 4). Adjusting M_0^* shifts the lines: decreasing M_0^* leads to a delayed onset of cover and vice versa. The former result arises because a lower M_0^* means that the sediment flux is conveyed through a smaller mass moving at a higher velocity. The original linear cover function (eq. 1) can be recovered from the exponential model with a high value of M_0^*, since the exponential term quickly becomes negligible with increasing $\overline{Q_s^*}$ and the linear term dominates (Fig. 4C). Note that for the linear (eq. 6) and the power law cases (eq. 10), high values of M_0^* may give $\overline{A^*} = 0$ for $\overline{Q_s^*} < 1$ (Fig. 4B,D), which is consistent with the concept of runaway alluviation. Using the beta distribution to describe P, a numerical solution is necessary, but a wide range of steady-state cover functions can be obtained (Fig. 5). By varying the value of M_0^*, an even wider range of behavior can be obtained.

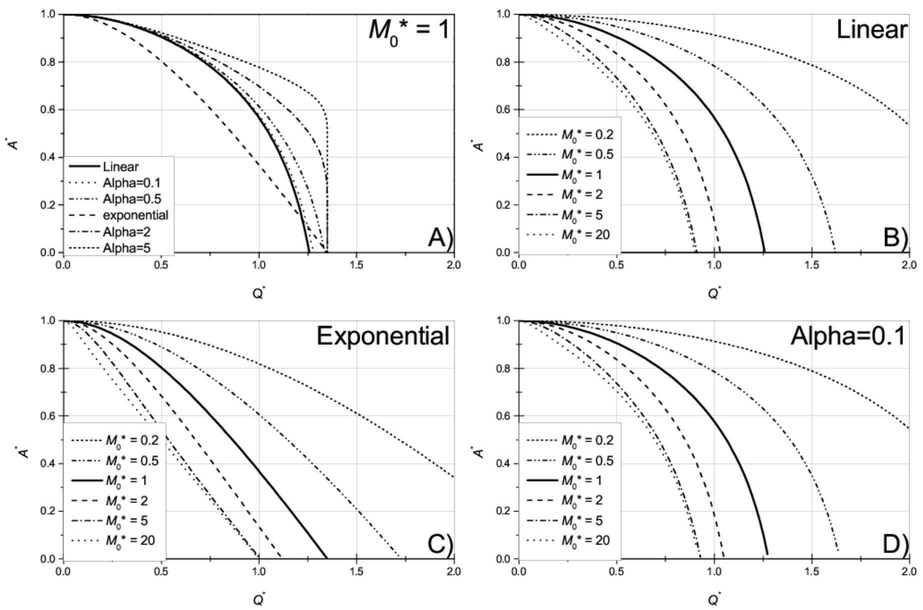

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Fig. 4: Analytical solutions at steady state for the exposed fraction of the bed (A*) as a function of relative sediment supply (Q*, cf. Fig. 1). A) Comparison of the different solutions, keeping M_0^* constant at 1. B) Varying M_0^* for the linear case (eq. 31). C) Varying M_0^* for the exponential case (eq. 30). D) Varying M_0^* for the power law case with α = 0.1 (eq. 32).

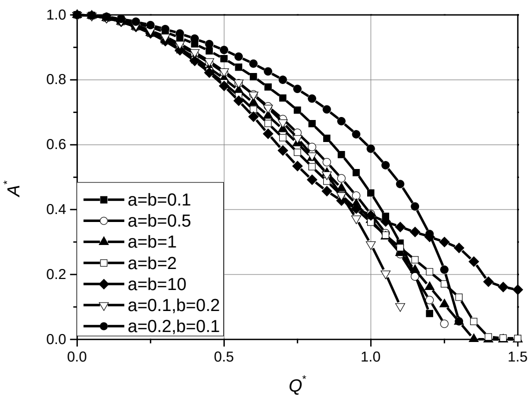

Fig. 5: Steady state solutions using the beta distribution to parameterize P (eq. 11) for a range of parameters a and b, and using M_0^* = 1 (cf. Fig. 2). The solutions were obtained by iterating the equations to a steady state, using initial conditions of A^* = 1 and M_m^* = M_s^* = 0.

The previous analysis shows that steady state cover is controlled by the characteristic dimensionless mass M_0^*, which is equal to the ratio of dimensionless transport capacity and particle speed (eq. 28). Converting to dimensional variables, we can write

$$M_0^* = \frac{q_t^*}{U^*} = \frac{q_t}{M_0 U}$$

(eq. 33)

The minimum mass necessary to completely cover the bed per unit area, M_0, can be estimated assuming a single layer of close-packed spherical grains residing on the bed (cf. Turowski, 2009), giving:

$$M_0 = \frac{\pi \rho_s D_{50}}{3\sqrt{3}}$$

(eq. 34)

Here, ρ_s is the sediment density and D_{50} is the median grain size. Fernandez-Luque and van Beek (1976) derived equations both for the transport capacity and the particle speed from flume experiments, using bed shear stress as a control parameter (see also Lajeunesse et al., 2010 and Meyer-Peter and Mueller, 1948 for similar equations).

$$q_t = 5.7 \frac{\rho_s \rho}{(\rho_s - \rho) g} \left(\frac{\tau}{\rho} - \frac{\tau_c}{\rho}\right)^{3/2}$$

(eq. 35)

$$U = 11.5 \left(\left(\frac{\tau}{\rho}\right)^{1/2} - 0.7 \left(\frac{\tau_c}{\rho}\right)^{1/2} \right)$$

(eq. 36)

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Here, τ_c is the critical bed shear stress for the onset of bedload motion, g is the acceleration due to gravity and ρ is the water density. Combining eqs. (34), (35) and (36) to get an equation for M_0^* gives:

$$M_0^* = \frac{3\sqrt{3}}{2\pi}\frac{(\theta-\theta_c)^{3/2}}{\theta^{1/2}-0.7\theta_c^{1/2}} = \frac{3\sqrt{3}\theta_c}{2\pi}\frac{(\theta/\theta_c-1)^{3/2}}{(\theta/\theta_c)^{1/2}-0.7}$$

(eq. 37)

Here, the Shields stress $\theta = \tau/(\rho_s - \rho)gD_{50}$, and θ_c is the corresponding critical Shields stress, and we approximated 5.7/11.5 = 0.496 with 1/2. At high θ, when the threshold can be neglected, eq. (37) reduces to a linear relationship between M_0^* and θ. Near the threshold, M_0^* is shifted to lower values as θ_c increases (Fig. 6). The systematic variation of U^* with the hydraulic driving conditions (eq. 36) implies that the cover function evolves differently in response to changes in sediment supply and transport capacity. For a first impression, by comparing equations (35) and (36), we assume that particle speed scales with transport capacity raised to the power of one third (Fig. 7).

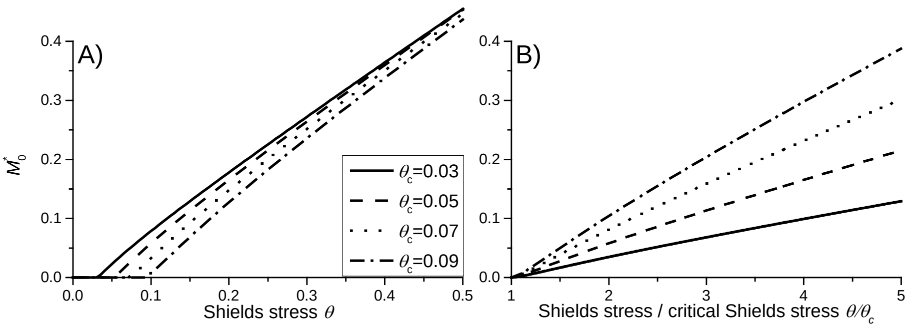

Fig. 6: The characteristic dimensionless mass M_0^* depicted as a function of A) the Shields stress and B) the ratio of Shields stress to critical Shields stress (eq. 37).

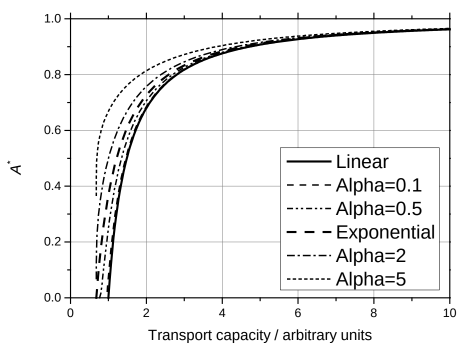

Fig. 7: Variation of the exposed bed fraction as a function of transport capacity, assuming that particle speed scales with transport capacity to the power of one third.

3.3 Temporal evolution of cover within a reach

3.3.1 System timescales

To calculate the temporal evolution of cover on the bed within a single reach, we solved the equations numerically for a section of the bed with homogenous conditions using a simple linear finite difference scheme. Then, the sediment input is a boundary condition, while sediment output,

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mobile and stationary sediment mass and the fraction of cover are output variables. In general, a change in sediment supply leads to a gradual adjustment of the output variables towards a new steady state (Fig. 8). Unfortunately, a general analytical solution is not possible, but a results can be obtained for the special case of $q_s^* = 0$. Such a situation is rare in nature, but could be easily created in flume experiments as a model test. Then, the time derivative of stationary mass is given by:

$$\frac{\partial M_S^*}{\partial t^*} = -\left(1 - e^{-M_S^*}\right) q_t^*$$

(eq. 38)

Using the exponential cover model (eq. 9), we obtain:

$$\frac{1}{A^*(1 - A^*)} \frac{\partial A^*}{\partial t^*} = q_t^*$$

(eq. 39)

Equation (39) is separable and can be integrated to obtain

$$\ln(A^*) - \ln(1 - A^*) = t^* q_t^* + C$$

(eq. 40)

Letting $A^*(t^*=0) = A^*_0$, where A^*_0 is the initial cover, the final equation is

$$\frac{1 - A^*}{1 - A_0^*} \frac{A_0^*}{A^*} = e^{-t^* q_t^*}$$

(eq. 41)

To clarify the characteristic time scale of the process, equation (41) can also be written in the form of a sigmoidal-type function:

$$A^* = \frac{1}{1 + \left(\frac{1 - A_0^*}{A_0^*}\right) e^{-t^* q_t^*}}$$

(eq. 42)

By making the parameters in the exponent on the right hand side of eq. (42) dimensional, we get:

$$t^* q_t^* = \frac{t}{T} \frac{T}{L M_0} q_t = \frac{t q_t}{L M_0}$$

(eq. 43)

which allows a characteristic system time scale T_E to be defined as

$$T_E = \frac{L M_0}{q_t}$$

(eq. 44)

Since this time scale is dependent on the transport capacity q_t, we can view it as a time scale associated with the entrainment of sediment from the bed (cf. eq. 20) – hence the subscript E on T_E. From eq. (42), the exposed bed fraction evolves in an asymptotic fashion towards equilibrium (Fig. 9). We can expect that there are other characteristic time scales for the system, for example associated with sediment deposition or downstream sediment evacuation.

We can make some further progress and define a more general system time scale by performing a perturbation analysis (Appendix A). For small perturbations in either q_s^* or q_t^*, we obtain an exponential term describing the transient evolution, which allows the definition of a system timescale T_S

$$\exp\left\{-\left(\overline{q_t^*}-\left(1-e^{-\overline{q_s^*}/\overline{U^*}}\right)\overline{q_s^*}\right)t^*\right\}=\exp\left\{-\frac{t}{T_S}\right\}$$

(eq. 45)

The characteristic system time scale can then be written as

$$T_S=\frac{LM_0}{\overline{q}_t\left(1-\left(1-e^{-\overline{q_s^*}/\overline{U^*}}\right)\frac{\overline{q}_s}{\overline{q}_t}\right)}=\frac{LM_0}{\overline{q}_t}e^{\overline{M_s^*}}$$

(eq. 46)

Note that for $q_s^* = 0$, eq. (46) reduces to eq. (44), as would be expected. Since $\overline{M_s^*}$ is directly related to steady state bed exposure $\overline{A^*}$, we can rewrite the equation, for example by assuming the exponential cover function (eq. 3), as

$$T_S=\frac{LM_0}{q_t\overline{A^*}}$$

(eq. 47)

Since bed cover is more easily measurable than the mass on the bed, eq. (47) can help to estimate system time scales in the field. Further, $\overline{A^*}$ varies between 0 and 1, which allows estimating a minimum system time using eq. (44). As $\overline{A^*}$ approaches zero, the system time diverges.

To illustrate these additional dependencies, we have calculated the time need to reach 99.9% (chosen due to the asymptotic behavior of the system) of total adjustment after a step change in transport stage, produced by varying particle speed U over a range of plausible values (Fig. 10). Response time decreases as particle speed increases. This reflects elevated downstream evacuation for higher particles speeds, resulting in a smaller mobile particle mass and thus higher entrainment and lower deposition rates. Response time also increases with increasing q_s/q_t. As the runs start with zero sediment cover, and the extent of cover increases with q_s/q_t, at higher q_s/q_t the adjusted cover takes longer to develop.

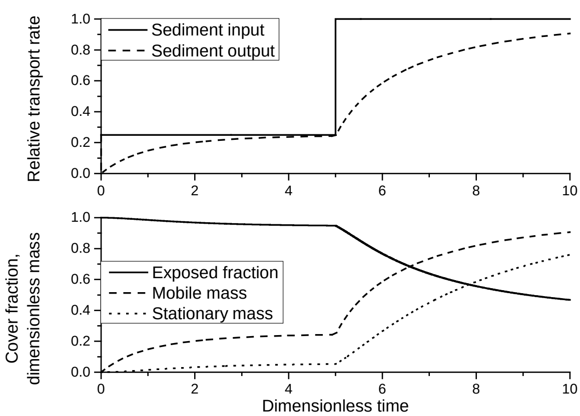

Fig. 8: Temporal evolution of cover for a simple case. Here, we used the exponential function for P (eq. 9) and $M_0^* = 1$. The initial values were $A^* = 1$, $M_m^* = M_s^* = 0$ and $q_s^* = 0.25$. Sediment supply was increased to $q_s^* = 1$ at $t^* = 5$.

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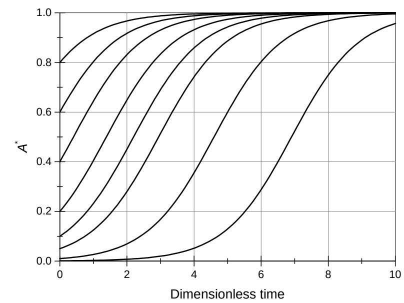

Fig. 9: Evolution of the exposed bed fraction (removal of sediment cover) over time starting with different initial values of bed exposure, for the special case q_s^* = 0 (eq. 41) and q_t^* = 1.

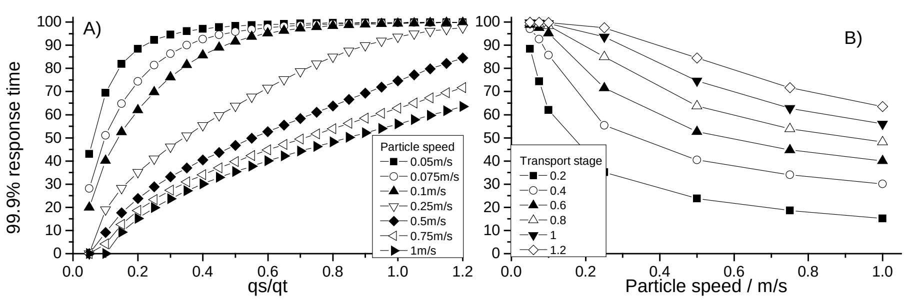

Fig. 10: Dimensionless time to reach 99.9% of the total adjustment in exposed area as a function of A) transport stage and B) particle speed. All simulation were started with A^* = 1 and M_m^* = M_s^* = 0.

3.3.2 Phase shift and gain in response to a cyclic perturbation

The perturbation analysis (Appendix A) gives some insight into the response of cover to cyclic sinusoidal perturbations. Let sediment supply be perturbed in a cyclic way described by an equation of the form

$$q_s^* = \overline{q_s^*} + \delta q_s^* = \overline{q_s^*} + d \sin\left(\frac{2\pi t}{p}\right)$$

(eq. 48)

Here, the overbar denotes the temporal average, δq_s^* is the time-dependent perturbation, d is the amplitude of the perturbation and p its period. A similar perturbation can be applied to the transport capacity (see Appendix A). The reaction of the stationary mass and therefore cover can then also be described by sinusoidal function of the form (Appendix A)

$$\delta M_s^* = G \sin\left(\frac{2\pi t}{p} + \varphi\right)$$

(eq. 49)

Here, δM_s^* is the perturbation of the stationary sediment mass around the temporal average, G is known as the gain, describing the amplitude response, and φ is the phase shift. If the gain is large, stationary mass reacts strongly to the perturbation; if it is small, the forcing does not leave a signal. The phase shift is negative if the response lags behind the forcing and positive if it leads. The phase shift can be written as

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$$\varphi = \tan^{-1}\left(-2\pi \frac{T_S}{p}\right)$$

(eq. 50)

The gain can be written as

$$G = \frac{p}{T_S} \frac{Kd}{\sqrt{\left(\frac{p}{T_S}\right)^2 + 4\pi^2}}$$

(eq. 51)

Here, d is the amplitude of the perturbation, and K is a function of the time-averaged values of q_s, q_t and U and differs for perturbations in transport capacity and sediment supply (see Appendix A). Thus, the system behavior can be interpreted as a function of the ratio of the period of perturbation p and the system time scale T_s. The period p is large if the forcing parameter, i.e., discharge or sediment supply, varies slowly and small when it varies quickly. According to eq. (50), the phase shift is equal to $-\pi/2$ for low values of p/T_s (quickly-varying forcing parameter), implying a substantial lag in the adjustment of cover. The phase shift tends to zero as p tends to infinity (Fig. 11). The gain varies approximately linearly with p/T_s for small p/T_s (quickly-varying forcing parameter), while it is approximately constant at a value of Kd for large p/T_s (slowly-varying forcing parameter) (eq. 51). Thus, if the forcing parameter varies slowly, cover adjustment keeps up at all times.

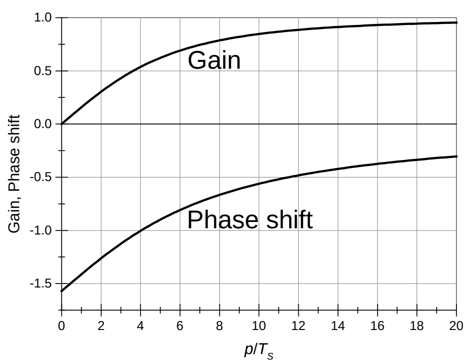

Fig. 11: Phase shift (eq. 50) and gain (eq. 51) as a function of the ratio of the period of perturbation period p and the system time scale T_s. For the calculation, the constant factor in the gain (Kd) was set equal to one.

3.3.3 A flood at the Erlenbach

To illustrate the magnitude of the timescales using real data, we use a flood dataset from the Erlenbach, a sediment transport observatory in the Swiss Prealps (e.g., Beer et al., 2015). There, near a discharge gauge, bedload transport rates are measured at 1-minute resolution using the Swiss Plate Geophone System, a highly developed and fully calibrated surrogate bedload measuring system (e.g., Rickenmann et al., 2012; Wyss et al. 2016). We use data from a flood on 20th June 2007 (Turowski et al., 2009) with highest peak discharge that has so far been observed at the Erlenbach. The meteorological conditions that triggered this flood and its geomorphic effects have been described in detail elsewhere (Molnar et al., 2010; Turowski et al., 2009). Although the Erlenbach does not have a bedrock bed in the sense that bedrock is exposed in the channel bed, the data provide a realistic natural time series of discharge and bedload transport over the course of a single event and are ideal for illustrating possible cover behavior.

Using a median grain size of 80 mm, a sediment density of 2650 kg/m^3 and a reach length of 50 m, we obtained $M_0 = 128$ kg/m^2. We calculated transport capacity using the equation of Fernandez Luque and van Beek (1976). However, it is known that this and similar equations strongly overestimate measured transport rates in streams such as the Erlenbach (e.g., Nitsche et al., 2011). Consequently, we rescaled by setting the ratio of bedload supply to capacity to one at the highest discharge. The exposed fraction was then calculated iteratively assuming $P = A^*$ (i.e., the exponential cover formulation). To estimate the period p, one needs to take the derivatives of eq. (48).

$$\frac{dq_s^*}{dt} = \frac{d\delta q_s^*}{dt} = \frac{2\pi d}{p}\cos\left(\frac{2\pi t}{p}\right)$$

(eq. 52)

Setting t = 0 for the time of interest, we can relate p to the local gradient in bedload supply, which can be measured from the data.

$$\frac{2\pi d}{p} = \frac{\Delta q_s^*}{\Delta t}$$

(eq. 52)

Assuming that all change in the response time is due to changes in the period (i.e., assuming constant amplitude, $d = 1$), we can obtain a conservative estimate of the range over which p varies over the course of an event.

$$p = 2\pi\frac{\Delta t}{\Delta q_s^*}$$

(eq. 52)

In the exemplary event, the evolution and final value of bed cover depends strongly on its initial value (Fig. 12), indicating that the adjustment is incomplete. The system timescale is generally larger than 1000s and is inversely related to discharge via the dependence on transport capacity. The p/T_s ratio varies around one, with low values at the beginning of the flood and large values in the waning hydrograph. Both the high system times and the smooth evolution of bed cover over the course of the flood imply that cover development cannot keep up with the variation in the forcing characteristics. This dynamic adjustment of cover, which can lag forcing processes, may thus play an important role in the dynamics of bedrock channels and probably needs to be taken into account in modelling exercises.

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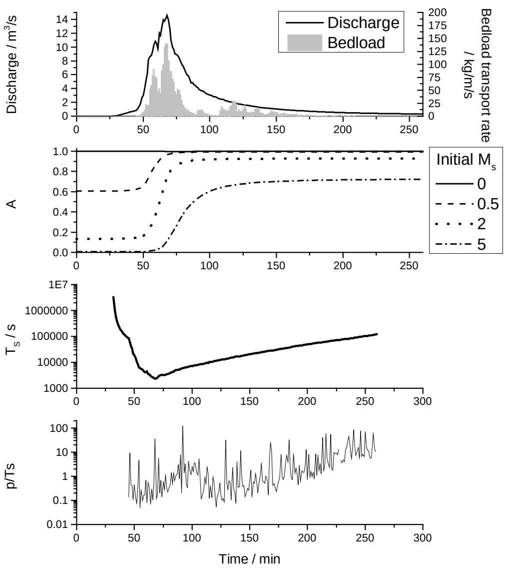

Fig. 12: Calculated evolution of cover during the largest event observed at the Erlenbach on 20th June 2007 (Turowski et al., 2009). Bedload transport rates were measured with the Swiss Plate geophone sensors calibrated with direct bedload samples (Rickenmann et al., 2012). The final fraction of exposed bedrock is strongly dependent on its initial value.

4. Discussion

4.1 Model formulation

In principle, the framework for the cover effect presented here allows the formulation of a general model for bedrock channel morphodynamics without the restrictions of previous models (e.g. Zhang et al., 2015). To achieve this, the dependency of P on various control parameters needs to be specified. In general, P should be controlled by local topography, grain size and shape, hydraulic forcing, and the amount of sediment already residing on the bed. Furthermore, the shape of the P function should also be affected by feedbacks between these properties, such as the development of sediment cover altering the local roughness and hence altering hydraulics and local transport capacity (Inoue et al., 2014; Johnson, 2014). Within the treatment presented here, we have explicitly accounted only for the impact of the amount of sediment already on the bed. However, all of the mentioned effects can be included implicitly by an appropriate choice of P. The exact relationships between, say, bed topography and P need to be mapped out experimentally (e.g., Inoue et al., 2014), with theoretical approaches also providing some direction (cf. Johnson, 2014; Zhang et al., 2015). Currently available experimental results (Chatanantavet and Parker, 2008; Finnegan et al., 2007; Hodge and Hoey, 2016; Inoue et al., 2014; Johnson and Whipple, 2007) cover only a small range of the possible parameter space and do not generally report all necessary parameters. Specifically the stationary mass of sediment residing on the bed is generally not reported and can be difficult to determine experimentally, but is necessary to determine P. Nevertheless, depending on the choice of P, our model can yield a wide range of cover functions that encompasses reported functions both from numerical modelling (e.g., Aubert et al., 2016; Hodge and Hoey, 2012; Johnson, 2014) and experiments (Chatanantavet and Parker, 2008; Inoue et al., 2014; Sklar and Dietrich, 2001).

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The dynamic model put forward here is a minimum first order formulation, and there are some obvious future alterations. We only take account of the static cover effect caused by immobile sediment on the bed. The dynamic cover effect, which arises when moving grains interact at high sediment concentration and thus reduce the number of impacts on the bed (Turowski et al., 2007), could in principle be included into the formulation, but would necessitate a second probability function specifically to describe this dynamics cover. It would also be possible to use different P-functions for entrainment and deposition, thus introducing hysteresis into cover development. Such hysteresis has been observed in experiments in which the equilibrium sediment cover was a function of the initial extent of sediment cover (Chatanantavet and Parker, 2008; Hodge and Hoey, 2012). Whether such alterations are necessary is best established with targeted laboratory experiments.

4.2 Comparison to previous modelling frameworks

We will briefly outline in this section the main differences to previous formulations of cover dynamics in bedrock channels. Thus, the novel aspects of our formulation and the respective advantages and disadvantages will become clear.

Aubert et al. (2015) coupled the movement of spherical particles to the simulation of a turbulent fluid and investigated how cover depended on transport capacity and supply. Similar to what is predicted by our analytical formulation, they found a range of cover function for various model set-ups, including linear and convex-up relationships (compare the results in Fig. 4 to their Fig. 15). Despite short-comings, Aubert et al. (2015) presented the so far most detailed physical simulations of bed cover formation and the correspondence between the predictions is encouraging.

Nelson and Seminara (2011, 2011) formulated a morphodynamic model for bedrock channels. They based their formulation on sediment concentration, which is in principle similar to our formulation based on mass. However, Nelson and Seminara (2011, 2012) did not distinguish between mobile and stationary sediment and linked local transport directly to sediment concentration. Further, a given mass can be distributed in multiple ways to achieve various degrees of cover, a fact that is quantified in our formulation by the probability parameter P. Nelson and Seminara (2011, 2012) assumed a direct correspondence between sediment concentration and degree of cover, which is equivalent to the linear cover assumption (eq. 7), with the associated problems outlined earlier. Practically, this implies that the grid size needs to be of the order of the grain size. Although different in various details, Inoue et al. (2016) have used essentially the same approach as Nelson and Seminar (2011, 2012) to link bedload concentration, transport and bed cover. Both of these models allow the 2D modelling of bedrock channel morphology. Although we have not fully developed such a model in the present paper, our model framework could easily be extended to 2D problems.

Zhang et al. (2015) formulated a bed cover model specifically for beds with macro-roughness. There, deposited sediment always fills topographic lows from their deepest positions, such that there is a reach-uniform sediment level. While the model is interesting and provides a fundamentally different approach to what is suggested here, its applicability is limited to very rough beds and the assumption of a sediment elevation that is independent of the position on the bed seems physically unrealistic. In principle, the probabilistic framework presented here should be able to deal with macro-rough beds as well and thus allows a more general treatment of the problem of bed cover.

Within this paper, we focused on the dynamics of bed cover, rather than modelling the dynamics of entire channels. The probabilistic formulation using the parameter P provides a flexible framework to connect the sediment mass residing on the bed with the exposed bedrock fraction. This particular element has not been treated in any of the previous models and could be easily implemented in other approaches dealing with sediment fluxes along and across the stream and the interaction with erosion and, over long time scales, channel morphology. However, it is as yet unclear how flow hydraulics, sediment properties and other conditions affect P and this should be investigated in targeted laboratory experiments. Nevertheless, the proposed formulation provides a framework in which data from various sources can be easily compared and discussed.

4.3 Further implications

Based on field data interpretation, Phillips and Jerolmack (2016) argued that bedrock rivers adjust such that, similar to alluvial channels, medium sized floods are most effective in transporting sediment, and that channel geometry therefore can quickly adjust their transport capacity to the applied load and therefore achieve grade (cf. Mackin, 1948). Contrary to the suggestion of Phillips and Jerolmack (2016) that this is achieved by changing channel morphologic parameters such as width, our model suggests that bed cover is adjusted. Furthermore changes in sediment cover can occur far more rapidly than morphological changes. In steady state, time derivatives need to be equal to zero to be equal to zero. Thus, entrainment equals deposition (eq. 16), implying that the downstream gradient in sediment transport rate is equal to zero (eq. 14). When sediment supply or transport capacity change, the exposed bedrock fraction can adjust to achieve a new steady state and a change of the channel geometry is unnecessary. Whether a steady state is achieved depends on the relative magnitude of the timescales of perturbation and cover adjustment (see section 3.2).

5. Conclusions

The probabilistic view put forward in this paper offers a framework into which diverse data on bed cover, whether obtained from field studies, laboratory experiments or numerical modeling, can be easily converted to be meaningfully compared. The conversion requires knowledge of the mass of sediment on the bed and the evolution of exposed fraction of the bed. Within the framework, individual data sets can be compared to the exponential benchmark and linear limit cases, enabling physical interpretation. Furthermore, the formulation allows the general dynamic sub-grid modelling of bed cover. Depending on the choice of P, the model yields a wide range of possible cover functions. Which of these functions are appropriate for natural rivers and how they vary with factors including topography needs to mapped out experimentally.

It needs to be noted here that the precise formulation of the entrainment and deposition functions also affects steady state cover relations. When calibrating P on data, it cannot always be decided whether a specific deviation from the benchmark case results from varying entrainment and deposition processes or from changes in the probability function driven for example by variations in roughness. For the prediction of the steady state cover relations and for the comparison of data sets, this should not matter, but the dynamic evolution of cover could be strongly affected.

The system timescale for cover adjustment is inversely related to transport capacity. This time scale can be long and in many realistic situations, cover cannot instantaneously adjust to changes in the forcing conditions. Thus, dynamic cover adjustment needs to be taken into account when modelling the long-term evolution of bedrock channels.

Our model formulation implies that bedrock channels adjust bed cover to achieve grade. Therefore, bedrock channel evolution is driven by two optimization principles. On short time scales, bed cover

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adjusts to match the sediment output of a reach to its input. Over long time scales, width and slope of the channel evolve to match long-term incision rate to tectonic uplift or base level lowering rates.

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Appendix A: Perturbation analysis

Here, we derive the effect of a small sinusoidal perturbation of the driving variables, namely sediment supply q_s^* and transport capacity q_t^*, on cover development. The perturbation of the driving variables can be written as

$$q_s^* = \overline{q_s^*} + \delta q_s^*$$

(eq. A1)

$$q_t^* = \overline{q_t^*} + \delta q_t^*$$

(eq. A2)

Here, the bar denotes the average of the quantity at steady state, while δq_s^* and δq_t^* denote the small perturbation. The exposed area can be similarly written as

$$A^* = \overline{A^*} + \delta A^*$$

(eq. A3)

Steady state cover is directly related to the mass on the bed M_s^* by eq. (3), which we can rewrite as

$$\frac{dA^*}{dt} = -P\frac{dM_s^*}{dt}$$

(eq. A4)

Substituting eq. (A3) and a similar equation for M_s^*,

$$M_s^* = \overline{M_s^*} + \delta M_s^*$$

(eq. A5)

we obtain

$$\frac{d\delta A^*}{dt} = -P\frac{d\delta M_s^*}{dt}$$

(eq. A6)

Here, the averaged terms drop out as they are independent of time. If P and the steady state solution for A^* are known, a direct relationship between A^* and M_s^* can be derived. For example, for the exponential cover model (eq. 2), substituting eqs. (A3) and (A5), we find

$$\overline{A^*} + \delta A^* = e^{-\overline{M_s^*} - \delta M_s^*} = e^{-\overline{M_s^*}} e^{-\delta M_s^*} = \overline{A^*} e^{-\delta M_s^*} \approx \overline{A^*}(1 - \delta M_s^*)$$

(eq. A7)

Here, since the δ variables are small, we approximated the exponential term using a Taylor expansion to first order. We obtain

$$\delta A^* = -\overline{A^*}\delta M_s^*$$

(eq. A8)

It is therefore sufficient to derive the perturbation solution for M_s^*, the time evolution of which is given by eq. (22). Eliminating M_m^* using eq. (24), we obtain

$$\frac{\partial M_s^*}{\partial t^*} = \left(1 - e^{-q_s^*/U^*}\right) q_s^* - \left(1 - e^{-M_s^*}\right) q_t^*$$

(eq. A9)

Perturbation of sediment supply

First, let's look at a perturbation of sediment supply q_s^*, while other parameters are held constant. Substituting eq. (A1) and (A5) into (A9), we obtain

$$\frac{\partial \delta M_s^*}{\partial t^*} = \left(1 - e^{-\left(\overline{q_s^*} + \delta q_s^*\right)/U^*}\right)\left(\overline{q_s^*} + \delta q_s^*\right) - \left(1 - e^{-\overline{M_s^*} - \delta M_s^*}\right) q_t^*$$

(eq. A10)

Again, since the δ variables are small, we can replace the relevant exponentials with Taylor expansion to first order:

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$$e^{-\delta q_s^*/U^*} \approx 1 - \frac{\delta q_s^*}{U^*}$$

(eq. A11)

A similar approximation applies for the exponential in M_s^*. Substituting eq. (A11) into eq. (A10), expanding the multiplicative terms, dropping terms of second order in the δ variables and rearranging, we get

$$\frac{\partial \delta M_s^*}{\partial t^*} = \delta q_s^* \left(1 - e^{-\overline{q_s^*}/U^*} + \frac{\overline{q_s^*}}{U^*} e^{-\overline{q_s^*}/U^*}\right) - \delta M_s^* \left(q_t^* - \left(1 - e^{-\overline{q_s^*}/U^*}\right)\overline{q_s^*}\right)$$

(eq. A12)

The perturbation is assumed to be sinusoidal

$$\delta q_s^* = d \sin\left(\frac{2\pi t}{p}\right)$$

(eq. A13)

Here, p is the period of the perturbation and d is its amplitude. Note that, to be consistent with the assumptions previously made, d needs to be small in comparison with the average sediment supply. Substituting, eq. (A12) can be integrated to obtain the solution

$$\delta M_s^* = G_{q_s^*} \sin\left(\frac{2\pi t}{P} + \varphi_{q_s^*}\right) + C\exp\left\{-\left(q_t^* - \left(1 - e^{-\overline{q_s^*}/U^*}\right)\overline{q_s^*}\right)\frac{t}{T}\right\}$$

where C is a constant of integration. The gain is given by

$$G_{q_s^*} = \frac{p}{T}\frac{\left(1 - e^{-\overline{q_s^*}/U^*} + \frac{\overline{q_s^*}}{U^*} e^{-\overline{q_s^*}/U^*}\right)d}{\sqrt{\left(q_t^* - \left(1 - e^{-\overline{q_s^*}/U^*}\right)\overline{q_s^*}\right)^2 \left(\frac{p}{T}\right)^2 + 4\pi^2}}$$

(eq. A14)

And the phase shift by

$$\varphi_{q_s^*} = \tan^{-1}\left[-\frac{2\pi}{\frac{p}{T}\left(q_t^* - \left(1 - e^{-\overline{q_s^*}/U^*}\right)\overline{q_s^*}\right)}\right]$$

(eq. A15)

Perturbation of transport capacity

The perturbation of the transport capacity q_t^* is a little more complicated, since both q_t^* and U^* are explicitly dependent on hydraulics (e.g., shear stress; see eqs. 43 and 44), and thus U^* is implicitly dependent on q_t^* and δq_t^*. To circumvent this problem, we expand the exponential term featuring $U^*(\delta q_t^*)$ in eq. (A9) using a Taylor series expansion around $\delta q_t^* = 0$.

$$\exp\left\{-\frac{q_s^*}{U^*(\delta q_t^*)}\right\} \approx \exp\left\{-\frac{q_s^*}{U^*(\delta q_t^* = 0)}\right\}\left[1 - \frac{q_s^*}{U^{*2}(\delta q_t^* = 0)}\frac{\partial U^*}{\partial \delta q_t^*}(\delta q_t^* = 0)\delta q_t^*\right]$$

(eq. A16)

Both U^* and its derivative are constants when evaluated at $\delta q_t^* = 0$. We can thus write

$$\exp\left\{-\frac{q_s^*}{U^*}\right\} = \exp\left\{-\frac{\overline{q_s^*}}{\overline{U^*}}\right\}\left[1 - \frac{\overline{q_s^*}}{\overline{U^*}^2}\overline{\left(\frac{\partial U^*}{\partial \delta q_t^*}\right)}\delta q_t^*\right] = [1 - C_0 \delta q_t^*] e^{-\overline{q_s^*}/\overline{U^*}}$$

(eq. A17)

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Here, C_0 is a constant. Proceeding as before by substituting eq. (A2), (A8) and (A17) into (A9), expanding exponential terms containing δ variables, dropping terms of second order in the δ variables and rearranging, we obtain:

$$\frac{\partial \delta M_s^*}{\partial t^*} = \left(B q_s^* e^{-q_s^*/\overline{U^*}} + e^{-\overline{M_s^*}} - 1 \right) \delta q_t^* - \delta M_s^* \overline{q_t^*} e^{-\overline{M_s^*}}$$

(eq. A18)

A sinusoidal perturbation of the form

$$\delta q_t^* = d \sin\left(\frac{2\pi t}{p}\right)$$

(eq. A19)

yields the solution

$$\delta M_s^* = G_{q_t^*} \sin\left(\frac{2\pi t}{P} + \varphi_{q_t^*}\right) + C\exp\left\{-\left(\overline{q_t^*} - \left(1 - e^{-q_s^*/\overline{U^*}}\right) q_s^*\right)\frac{t}{p}\right\}\left\{-\left(\overline{q_t^*} - \left(1 - e^{-q_s^*/\overline{U^*}}\right) q_s^*\right)\frac{t}{T}\right\}$$

with

$$G_{q_t^*} = \frac{p}{T} \frac{\left(\frac{q_s^{*2}}{\overline{U^*}^2}\overline{\left(\frac{\partial U^*}{\partial \delta q_t^*}\right)} e^{-q_s^*/\overline{U^*}} - \left(1 - e^{-q_s^*/\overline{U^*}}\right)\frac{\underline{\underline{q_s^*}}}{q_t^*}\right) d}{\sqrt{\overline{q_t^*}^2 \left(\frac{p}{T}\right)^2 \left(1 - \left(1 - e^{-q_s^*/U^*}\right)\frac{\underline{\underline{q_s^*}}}{q_t^*}\right)^2 + 4\pi^2}}$$

(eq. A20)

and

$$\varphi = \tan^{-1}\left(- \frac{2\pi}{\frac{p}{T}\left(\overline{q_t^*} - \left(1 - e^{-q_s^*/\overline{U^*}}\right) q_s^*\right)} \right)$$

(eq. A21)

Summary

Using the system timescale T_S, the phase shift and gain can be generally rewritten as

$$\varphi = \tan^{-1}\left(-2\pi \frac{T_S}{p}\right)$$

(eq. A22)

$$G = \frac{p}{T_S} \frac{Kd}{\sqrt{\left(\frac{p}{T_S}\right)^2 + 4\pi^2}}$$

(eq. A23)

Here, K differs for perturbations in sediment supply and transport capacity, given by the equations

$$K_{q_s^*} = 1 - e^{-\overline{q_s^*}/U^*} + \frac{\overline{q_s^*}}{U^*} e^{-\overline{q_s^*}/U^*}$$

(eq. A24)

$$K_{q_t^*} = \frac{q_s^{*2}}{\overline{U^*}^2}\overline{\left(\frac{\partial U^*}{\partial \delta q_t^*}\right)} e^{-q_s^*/\overline{U^*}} - \left(1 - e^{-q_s^*/\overline{U^*}}\right)\frac{\underline{\underline{q_s^*}}}{q_t^*}$$

(eq. A25)

Notation

Overbars denote time-averaged quantities.

a	Shape parameter in the regularized incomplete Beta function.
A^*	Fraction of exposed (uncovered) bed area.
b	Shape parameter in the regularized incomplete Beta function.
B	Regularized incomplete Beta function.
C	Constant of integration.
C_0	Constant [m^2s/kg].
d	Amplitude of perturbation [kg/m^2s].
D	Sediment deposition rate per bed area [kg/m^2s].
D^*	Dimensionless sediment deposition rate.
D_{50}	Median grain size [m].
e	Base of the natural logarithm.
E	Sediment entrainment rate per bed area [kg/m^2s].
E^*	Dimensionless sediment entrainment rate.
E_{max}	Maximal possible dimensionless sediment entrainment rate.
g	Acceleration due to gravity [m/s^2].
G	Gain [kg/m^2s].
I	Non-dimensional incision rate.
k	Probability of sediment deposition on uncovered parts of the bed, linear implementation.
k_I	Non-dimensional erodibility.
K	Parameter in the gain equation.
L	Characteristic length scale [m].
M_0	Minimum mass per area necessary to cover the bed [kg/m^2].
M_0^*	Dimensionless characteristic sediment mass.
M_m	Mobile sediment mass [kg/m^2].
M_m^*	Dimensionless mobile sediment mass.
M_s	Stationary sediment mass [kg/m^2].
M_s^*	Dimensionless stationary sediment mass.
p	Period of perturbation [s].
P	Probability of sediment deposition on uncovered parts of the bed.
q_s	Mass sediment transport rate per unit width [kg/ms].
q_s^*	Dimensionless sediment transport rate.
q_t	Mass sediment transport capacity per unit width [kg/ms].
q_t^*	Dimensionless transport capacity.
Q_s^*	Relative sediment supply; sediment transport rate over transport capacity.
Q_t	Mass sediment transport capacity [kg/s].
t	Time variable [s].
t^*	Dimensionless time.
T	Characteristic time scale [s].
T_E	Characteristic time scale for sediment entrainment [s].
T_S	Characteristic system time scale [s].
U	Sediment speed [m/s].
U^*	Dimensionless sediment speed.
x	Dimensional streamwise spatial coordinate [m].

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x^*	Dimensionless streamwise spatial coordinate.
y	Dummy variable.
α	Exponent.
γ	Fraction of pore space in the sediment.
δ	denotes time-varying component.
θ	Shields stress.
θ_c	Critical Shields stress.
ρ	Density of water [kg/m^3].
ρ_s	Density of sediment [kg/m^3].
τ	Bed shear stress [N/m^2].
τ_c	Critical bed shear stress at the onset of bedload motion [N/m^2].

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